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SCATTERING OF REGIONAL P_n BY MOHO TOPOGRAPHY

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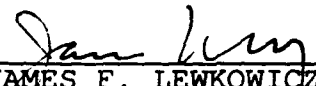
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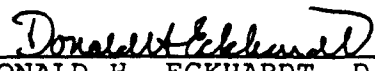
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19. ABSTRACT The often observed relatively large amplitudes in the later part of the P_n signal cannot be explained with the traditional interpretation of P_n in 1-D crust-mantle models. To determine the cause of these characteristics, we have analyzed in some detail the NORESS array records of P_n from a suite of quarry blasts in S.W. Norway. Application of wide-band frequency-wavenumber analysis to these records confirms that there is a slowness and azimuth anomaly associated with the dominant part of the wavetrain and that it is confined to a particular frequency range. Moreover, the scattering source of the anomaly is determined to be at Moho depth, which is consistent with the concept of scattering by topographic relief. We demonstrate the viability of this concept by means of numerical experiments, showing that for realistic models of topography, the energy flux of scattered P can dominate the specular flux (i.e., the flux in the direction defined by the ray crossing a plane interface) for incidence angles approaching the critical angle of P_n . Since the effects appear to be systematic, we have the possibility to calibrate the P_n parameters for event location and velocity determination purposes.					
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Preface

Under Contract No. F49620-C-89-0038, NTNF/NORSAR is conducting research within a wide range of subjects relevant to seismic monitoring. The emphasis of the research program is on developing and assessing methods for processing of data recorded by networks of small-aperture arrays and 3-component stations, for events both at regional and teleseismic distances. In addition, more general seismological research topics are addressed.

Each quarterly technical report under this contract presents one or several separate investigations addressing specific problems within the scope of the statement of work. Summaries of the research efforts within the program as a whole are given in annual technical reports.

This Scientific Report No. 8 presents a manuscript entitled "Scattering of regional Pn by Moho topography", by T. Kværna and D.J. Doornbos.

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Introduction

The P_n is crucial for the detection and location of regional events. It is therefore important to determine its characteristic properties. The properties of P_n are well understood in one-dimensional crust-mantle models: P_n is an ordinary head wave associated with the Moho in models of uniform plane layers, and P_n can be interpreted as a sum of whispering gallery waves forming a so-called interference head wave in models of spherical layers and/or models having a positive velocity gradient below the Moho [Menke and Richards, 1980]. This model can explain the relatively long duration and high frequencies of P_n at teleseismic distances. However, it cannot explain the relatively large amplitudes often observed in the later part of the coda. Similar characteristics are also observed at regional distances.

In this study we apply wide-band frequency-wavenumber analysis to regional P_n at the NORESS array in southern Norway. The results suggest a scattering model for the propagation of P_n . Scattering due to surface topography has been invoked earlier to explain the often significant perturbations of three-component data at sites within the NORESS array [Ødegaard et al., 1990]. It has also been demonstrated that local surface scattering may be responsible for the relatively large coda amplitudes of teleseismic P waves [Gupta et al., 1990, Hedlin et al., 1990]; this is in accord with earlier theoretical results that the incident P waves are effectively scattered into Rayleigh waves by surface obstacles [e.g., Hudson and Boore, 1980]. However this particular mechanism does not explain the characteristics of regional P_n at NORESS.

The data

We have analyzed in some detail the NORESS records of P_n from a suite of six quarry blasts at a dam construction site, near Blåsjø in S.W. Norway. The site is

about 300 km from NORESS in an azimuth direction of 240° . Event information is given in Table 1. A typical record of P_n from these events is shown in the top panel of Figure 1. It can be seen that the first arrival is relatively weak. In fact, this arrival is easily missed for small events, as P_n is often associated with the dominant part of the wavetrain.

We have applied wide-band frequency-wavenumber analysis [Kværna and Doornbos, 1986] to all six events in Table 1. It is useful to display the results of this analysis both as a function of frequency and as a function of time. Estimates of slowness and azimuth as a function of time have been obtained by applying a moving window to the records; we used a window length of 0.75 s and a time step of 0.2 s. The relatively short window length requires that the array data be preshifted in accordance with the azimuth to the source and a specified apparent velocity; we used 8.2 km/s. Results for the frequency band 2–4 Hz are summarized in the two lower panels of Figure 1, in the form of slowness and azimuth as a function of time. These results represent an average over all six events.

The results show that the slowness (0.128 s/km) and azimuth (242°) of the first arrival are consistent with P_n in a one-dimensional crust-mantle model. The anomaly (slowness 0.138 s/km, azimuth 227°) is related to the dominant part of the wavetrain that is being delayed 0.5–0.6 s relative to the first arrival. This is a consistent feature for all events. Slowness and azimuth estimates as a function of frequency are summarized in Figure 2, showing that the anomaly is related to the frequency range 2–4 Hz; this is also the range where the signal has its maximum energy.

These results have led us to consider a scattering mechanism for the generation of the anomalous wavetrain. Scattering has previously been postulated at a wide range of depths in the crust and upper mantle; one especially expects this phenomenon to occur near the major discontinuities, near the Earth's surface and connected with major geological features. However, the observational results described above enable us to constrain the scatterer location to a considerable degree. The wavenumber solutions show that the wavefront is in fact very nearly plane. This implies that the anomaly cannot be generated near the surface since this would significantly perturb the wavefront. The observational results thus suggest that these " P_n " waves are actually the result of scattering at depth.

Ray-tracing backward in the direction determined by the measured slowness and azimuth of the anomalous main pulse leads to the result that the observed 0.5–0.6 s time delay relative to the first arrival is explained by a scattering source within the depth range of the Moho (30–40 km). We have traced the rays in a standard crustal and upper mantle model for southern Norway. Another result of this procedure is that scattering in the upper mantle far below the Moho would lead to a time residue significantly exceeding the observed 0.5–0.6 s. The Moho depth in the area is usually taken to be about 35 km [Berteussen, 1977]. Hence the observations are consistent with scattering by topographic relief of the Moho. An interesting geological feature is that the inferred location of the proposed topographic feature coincides with the border of the Oslo Graben (Figure 3).

Scattering by Moho topography

Scattering will of course affect all waves interacting with a rough Moho discontinuity. However, the scattered waves usually arrive in the coda of a relatively strong primary wave. In contrast, the first arriving P_n is relatively weak due to the small coefficient of refraction through the Moho, and scattering due to topography of the

boundary may dominate the wavetrain.

To illustrate these concepts, we have calculated generalized transmission coefficients for a rough Moho, using a recursion method for obtaining these coefficients in wavenumber space [Doornbos, 1988]. From the generalized coefficients we can calculate the energy flux for any wave type. Figure 4 shows the energy flux of P transmitted upward through the boundary, as a function of slowness of the incident P below. The boundary topography is characterized by a correlation length of 5.6 km and an average height of 1 km, and the wave frequency is 3 Hz. Two modes of scattering are shown: (1) The specular flux E^o in the direction defined by the plane wave-plane interface concept. The specular flux through a rough interface is reduced with respect to the flux through a plane. (2) The diffuse flux E^{sc} due to multiple scattering in all upward directions. E^{sc} does not exist for a plane interface. The figure illustrates well the sharp increase in the ratio E^{sc}/E^o as the slowness approaches the critical value corresponding to P_n , thus supporting in a qualitative way the scattering model for propagation of this wave. Within the context of this model, a likely interpretation of the frequency dependent slowness estimates, illustrated in Figure 2, is the well-known fact that the characteristic dimension of the boundary perturbation determines the frequency dependent radiation pattern of the scattering. It should be noted that E^o and E^{sc} are not observable parameters. What can be inferred is the flux at a receiver location on the surface. To model this we need to know the areal extent of topography, and the decay factor accounting for attenuation of P_n along the Moho. Clearly these parameters are presently only poorly constrained. Nevertheless, the present results suggest that the scattering model of P_n wave propagation is viable, and consequently that a careful calibration is needed before using this phase for event location and velocity determination purposes.

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Tables

Table 1. Event data for Blåsjø events

Event	Date	Origin time	Magnitude
1	06/04/85	03 13 59	2.3
2	06/07/85	14 42 39	2.0
3	06/28/85	15 42 12	2.0
4	07/05/85	03 59 29	2.3
5	07/16/85	17 33 12	2.5
6	07/29/85	13 55 18	2.8

Blåsjø coordinates: 59.3N, 6.95E.

NORESS coordinates: 60.74N, 11.54E

FIGURE CAPTIONS

Fig. 1. NORESS record of P_n from a quarry blast in the Blåsjø area. The top trace is a single channel seismogram, bandpass filtered between 2 and 4 Hz. Average array estimates of azimuth and slowness as a function of time are shown below. The error bars correspond to one standard deviation and the symbol size on the plot indicates the average relative signal power.

Fig. 2. Azimuth and slowness estimates of P_n as a function of frequency, for six events in the Blåsjø area. A 3 second time window was analyzed for each event, the frequency bandwidth was 0.8 Hz and the frequency step were 0.4 Hz. The symbol size indicates the relative signal power.

Fig. 3. Map showing the locations of Blåsjø and NORESS, as well as the surface projection of the presumed scattering region. The boundary of the Oslo Graben is indicated.

Fig. 4. Energy flux at 3 Hz through a rough solid-solid interface with the velocity-density structure $v_p^{+/-} = 6.8/8.1$ km/s, $v_s^{+/-} = 3.7/4.5$ km/s, $\rho^{+/-} = 2.9/3.4$ g/cm³. The boundary roughness is characterized by an average height of 1 km and a correlation length of 5.6 km. The scattered flux is upward through the boundary. Solid line: specular direction. Dashed line: integrated diffuse flux.

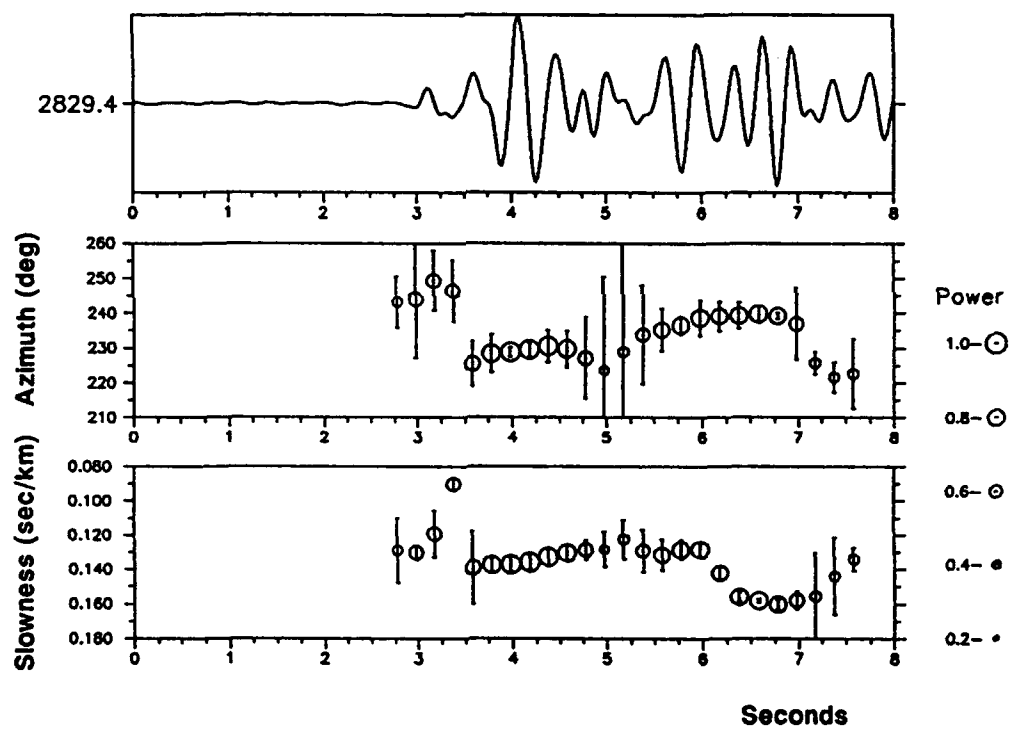


FIGURE 1

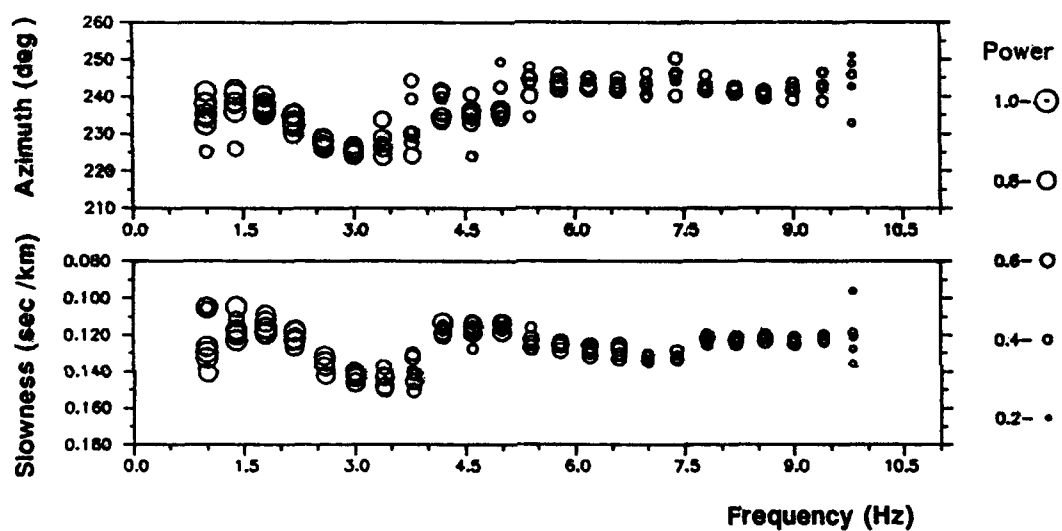
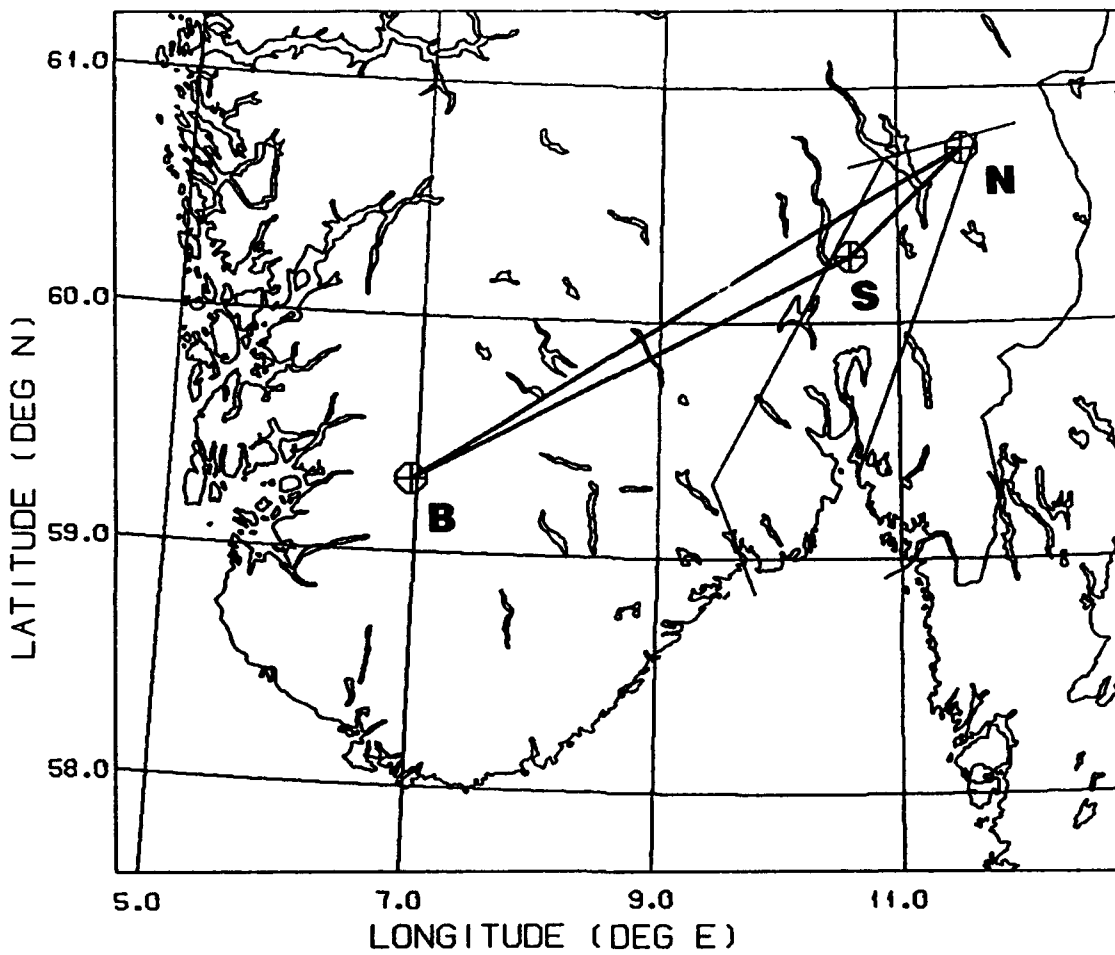


FIGURE 2



N - Noress
S - Scattering region
B - Blåsjø

FIGURE 3

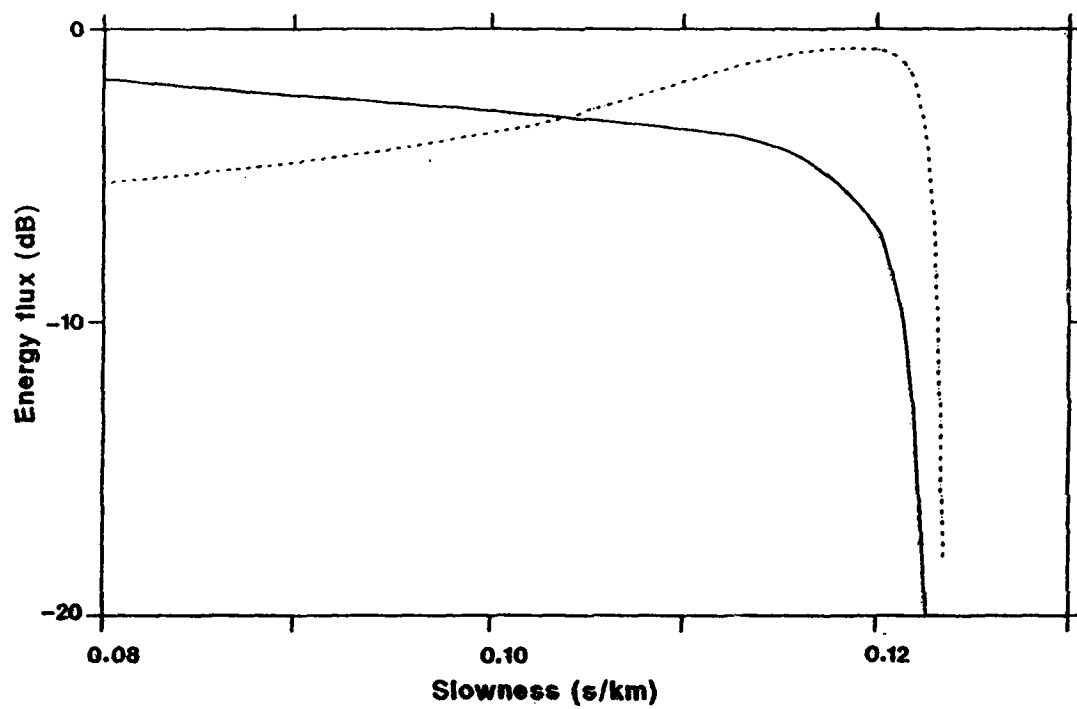


FIGURE 4

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